Modeling Flow and Chemical Quality Changes in an Irrigated Stream-Aquifer System

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Salinity increases in groundwater and surface water in the Arkansas River valley of southeastern Colorado are primarily related to irrigation practices. A digital computer model was developed to predict changes in dissolved solid concentration in response to spatially and temporally varying hydrologic stresses. The equations that describe the transient flow of groundwater and the transport and dispersion of dissolved chemical constituents were solved numerically. The model simulated flow as well as changes in water quality for both the stream and the aquifer. Detailed field measurements made for a 1-yr period in an 11-mi reach of the valley between La Junta and the Bent-Otero county line were used to verify and calibrate the model. Measured water levels varied by an average of about 3 ft during the study period, and calculated water table elevations in the aquifer were within 1 ft of the observed values approximately 90% of the time. The specific conductances of water samples from five wells in one well field had a standard deviation of about 10% of the mean. Dissolved solid concentrations calculated by the model were within 10% of the observed values for both the aquifer and the stream approximately 80% of the time.

Groundwater and surface water are interrelated in stream-aquifer systems in which groundwater in the floodplain alluvium is in hydraulic connection with the stream. In the arid to semiarid west the fertile floodplain soils are commonly irrigated with both diverted surface water and pumped shallow groundwater. Much of the applied irrigation water is lost by evapotranspiration, but some applied water recharges the alluvial aquifer and provides return flow to the stream.

Dissolved solids become concentrated in the recharged water because evapotranspiration consumes some of the applied water but has little effect on the total weight of chemical constituents dissolved in the water. The down-valley reuse of water causes a buildup of salts to levels that may approach concentrations intolerable to many crops. In several places this effect is so pronounced that the quality of water, rather than the amount available, restricts water use.

The return flow of irrigation water is sometimes the main cause of observed increases in salinity of river water [e.g., *Gordon*, 1966]. There is not at present a complete understanding of the relationship between irrigation practices and water quality variations. Numerous chemical and physical factors affect water quality, and it is difficult to isolate the influence of any one.

The maximum beneficial use of the total available water resource in a stream-aquifer system comes only through the conjunctive use of surface water and groundwater [*Young and Bredehoeft*, 1972]. But the conjunctive use can only be optimized if the interrelation between surface water and groundwater is understood and responses to physical and chemical stresses can be predicted. Successful water management thus depends largely upon the planners' knowledge of the water resource and the ability to predict both hydrologic and chemical effects over time of a complex set of dynamic stresses. Consequently, an accurate hydrologic and water quality simulation model of the stream-aquifer system is a desirable management tool for predicting responses and optimizing the development and use of the total resource. Optimization in turn can result in improved crop productivity.

With the advent of high-speed digital computers, numerical methods have been successfully applied to simulating groundwater flow and its relation to surface water. There are several numerical methods of modeling available, and all are based upon a basic knowledge of the physical properties and boundaries of the system. However, in the past, little success has been achieved in the modeling of concurrent water quality changes.

PURPOSE OF INVESTIGATION

The primary objective of this investigation was to develop, verify, and demonstrate the applicability of a digital computer model to simulate the hydrologic and dissolved solid concentration variations that occur in an irrigated stream-aquifer system in a semiarid climate. The model would then be used as an aid in interpreting and defining the relation between hydrologic stresses (such as pumpage, irrigation, and changes in streamflow) and the quality of water in the stream and aquifer. Furthermore, an assessment would be made of both the data requirements and the reliability of the model.

SELECTION OF STUDY AREA

The area selected for detailed investigation was an 11-mi reach of the Arkansas River valley in southeastern Colorado between La Junta and the Bent-Otero county line (Figure 1). This area was chosen primarily because detailed hydrogeologic data were available from previous studies. Also the hydrogeologic framework and water use patterns within the study reach are representative of most of the Arkansas River valley in southeastern Colorado and in fact are typical of many other irrigated river valleys in semiarid to arid areas. Furthermore, the area is supplied by only one major canal system, and tributary inflow was believed to be negligible.

A downstream increase in dissolved solid concentration in both surface water and groundwater has long been noted in the Arkansas River valley. Data indicate that the concentration of dissolved solids increases from less than 500 mg/l near Pueblo to more than 4000 mg/l at the Kansas State line [*Odell et al.*, 1964]. This increase is attributed primarily to return flow of irrigation water.

DESCRIPTION OF STUDY REACH

In the 11-mi study reach of the Arkansas River valley between La Junta and the Bent-Otero county line the alluvial

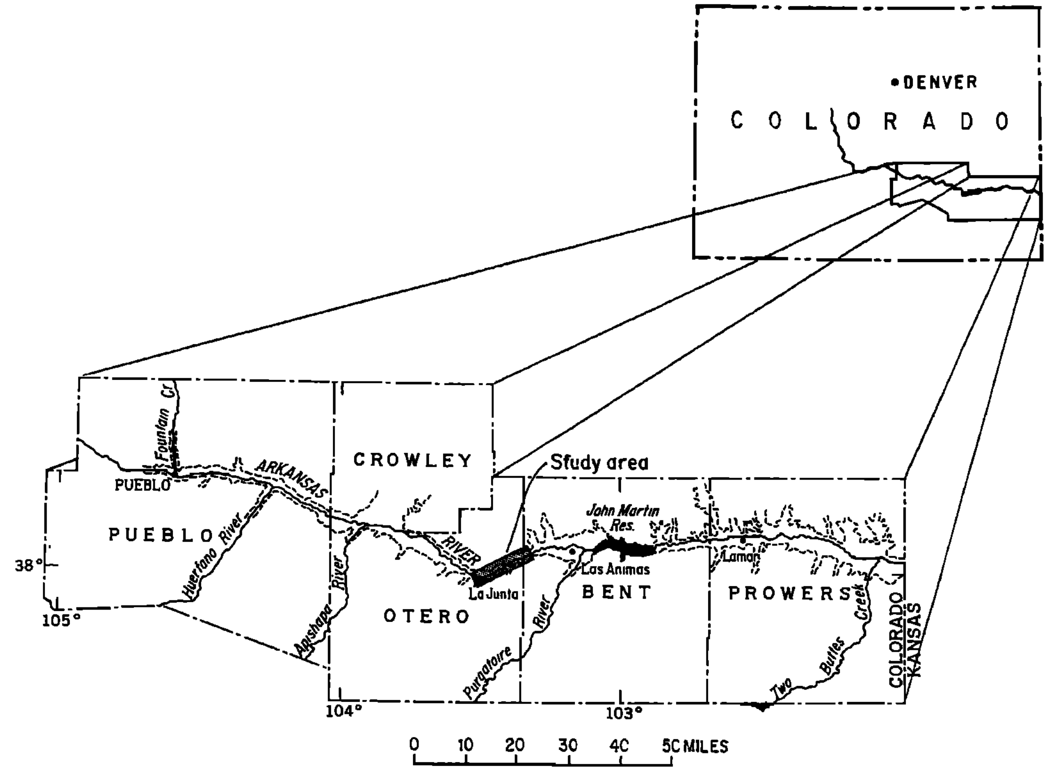

Fig. 1. Index map showing location of study area in southeastern Colorado.

valley is about 1½ mi wide and occupies an area of about 17 mi². The climate is semiarid, the average annual precipitation being 13 in. Two thirds of the annual precipitation occurs between May and September, primarily during brief but intense thunderstorms. The average annual temperature is 50°F, and the mean monthly average temperature ranges from about 30°F in January to 78°F in July.

The average discharge of the Arkansas River at La Junta is 260 ft³/s [*U.S. Geological Survey*, 1969, p. 81]. Highest flows normally occur between May and September. During 1960–1970, recorded daily mean discharges have ranged from less than 1 to about 20,000 ft³/s. A major diversion from the river into the Fort Lyon Canal is made west of La Junta just beyond the study area. Several minor tributaries and discharge from the La Junta municipal sewage treatment plant enter the river within the study reach.

The economy of the area is primarily agricultural; alfalfa, corn, and sugar beets are the main crops. Substantial irrigation is required during the growing season, and many farmers depend on both surface water from the canal and groundwater pumped from irrigation wells.

Previous Work in Study Area

The geology and occurrence of groundwater in the study area were described by *Weist* [1965]. Records of wells (including locations, estimated yields, depths to water, available well logs, and other data) were presented by *Weist* [1962] and by *Major et al.* [1970]. Detailed hydrologic data (aquifer boundaries, transmissivities, specific yield, and saturated thicknesses) were compiled for the study area in Otero County by R. T. Hurr and J. E. Moore (personal communication, 1971) and in adjacent Bent County by *Hurr and Moore* [1972].

These detailed data were used to construct an electrical analog model of the Arkansas River valley in southeastern Colorado to predict water level behavior in the aquifer in response to hydrologic stresses. *Moore and Wood* [1967] discuss the construction, calibration, and use of the analog model in the study area. *Longenbaugh* [1967] used a digital computer to develop a hydrologic simulation model of a 25-mi reach that included the present study area. However, his model did not consider water quality changes.

Procedure of the Investigation

The study was designed to simulate a 1-yr period of record (March 1971 through February 1972) that would include one complete irrigation season. Variations of hydrologic and water quality parameters with time were observed and simulated on a monthly basis. Several supplementary gages, water quality monitoring stations, and observation wells were constructed for this study to provide the necessary data for input to the model. Detailed data requirements for this study or equivalent studies are discussed in a following section.

A working simulation model of groundwater hydraulics is a prerequisite for a successful water quality model, and a preliminary groundwater flow model was developed for this study while data were still being collected for the water quality study. Monthly data were available for the 25-mi reach of the Arkansas River from La Junta to Las Animas for a 3-yr period (1966–1968). The specific details of this preliminary hydrologic model study are not included in this report, but the results led to the following conclusions:

1. The digital modeling technique used can satisfactorily simulate the hydrology of the study reach and reproduce observed changes in both water table elevations and streamflow in response to hydrologic stresses.

2. Hydrologic data obtained from previous studies describe the aquifer properties and are of sufficient accuracy.

3. Recharge to the aquifer from irrigation can be estimated as a function of applied water and monthly potential evapotranspiration computed by using the method described by *Blaney and Criddle* [1962].

4. Spatial variations in recharge should be considered, improvements in measuring water applied for irrigation thereby being necessitated.

5. An accurate water table map is required.

6. The model responses appeared sensitive to adjustments of potential evapotranspiration, canal leakage, and the way in which the river itself is approximated in the model but were relatively insensitive to variations in precipitation, well pumpage, and underflow through the alluvium.

DIGITAL SIMULATION MODEL

The hydraulics of a stream-aquifer system may in general be simulated mathematically by using analytical methods, analog models, or digital models. Standard analytical methods can be used if the system is highly idealized but are not readily suited for use with complex boundary conditions, spatially varying aquifer parameters, and complex pumping schemes such as those that exist in the Arkansas River valley. Electric analog models are well suited to such hydrologic problems but cannot readily be coupled to a water quality model. Digital computer models are also well suited for the solution of such problems and can readily be programed to simulate water quality changes.

The computer model used in this investigation solves a set of two simultaneous partial differential equations. One equation is the equation of flow that describes the head distribution in the aquifer. If the head distribution is given, the flow can be calculated by using Darcy's law. The second partial differential equation is the transport equation (dispersion equation) that describes the chemical concentration in the system.

Flow equation. By following the derivations of *Pinder and Bredehoeft* [1968], *Pinder* [1970], and *Bredehoeft and Pinder* [1973] the differential equation describing the nonsteady two-dimensional flow of a homogeneous compressible fluid in a nonhomogeneous anisotropic aquifer may be written

$$\frac{\partial}{\partial x_i}\left(T_{ij}\frac{\partial h}{\partial x_j}\right) = S\frac{\partial h}{\partial t} + W(x, y, t) \qquad (1)$$

where T_{ij} is the transmissivity tensor, L^2/T; h is the hydraulic head, L; S is the storage coefficient (dimensionless); t is time, T; and W is the volume flux per unit area, L/T. In the stream-aquifer system being studied the flux term may represent pumpage from or recharge to the aquifer or stream gains and losses. Thus $W(x, y, t)$ may be expressed as

$$W(x, y, t) = Q(x, y) + \frac{K_z}{m}(H_r - h) \qquad (2)$$

where Q is the withdrawal or recharge, L/T; K_z is the vertical hydraulic conductivity of the stream bed, L/T; m is the thickness of the stream bed, L; and H_r is the hydraulic head in the river, L.

If the coordinate axes are aligned with the principal directions of the transmissivity tensor, (1) may be approximated by the following finite difference equation:

$$\begin{aligned}
&T_{xx[i-(1/2),j]}\left[\frac{h_{i-1,j,k} - h_{i,j,k}}{(\Delta x)^2}\right] \\
&\quad + T_{xx[i+(1/2),j]}\left[\frac{h_{i+1,j,k} - h_{i,j,k}}{(\Delta x)^2}\right] \\
&\quad + T_{yy[i,j-(1/2)]}\left[\frac{h_{i,j-1,k} - h_{i,j,k}}{(\Delta y)^2}\right] \\
&\quad + T_{yy[i,j+(1/2)]}\left[\frac{h_{i,j+1,k} - h_{i,j,k}}{(\Delta y)^2}\right] \\
&\quad = S\left[\frac{h_{i,j,k} - h_{i,j,k-1}}{\Delta t}\right] \\
&\quad + \frac{q_{w(i,j)}}{\Delta x \Delta y} - \frac{K_z}{m}[H_{r(i,j)} - h_{i,j,k-1}] \qquad (3)
\end{aligned}$$

where $q_{w(i,j)}$ is the rate of withdrawal or recharge at the (i, j) node, L^3/T.

Analysis of the complex nonhomogeneous stream-aquifer system is accomplished by subdividing it into a large number of relatively small rectangular cells, which constitute a finite difference grid. Figure 2 is a map of the study reach showing the rectangular uniformly spaced block-centered finite difference grid that was imposed upon the system. The grid consists of 20 rows and 44 columns. Each cell has dimensions of 660 by 1320 ft.

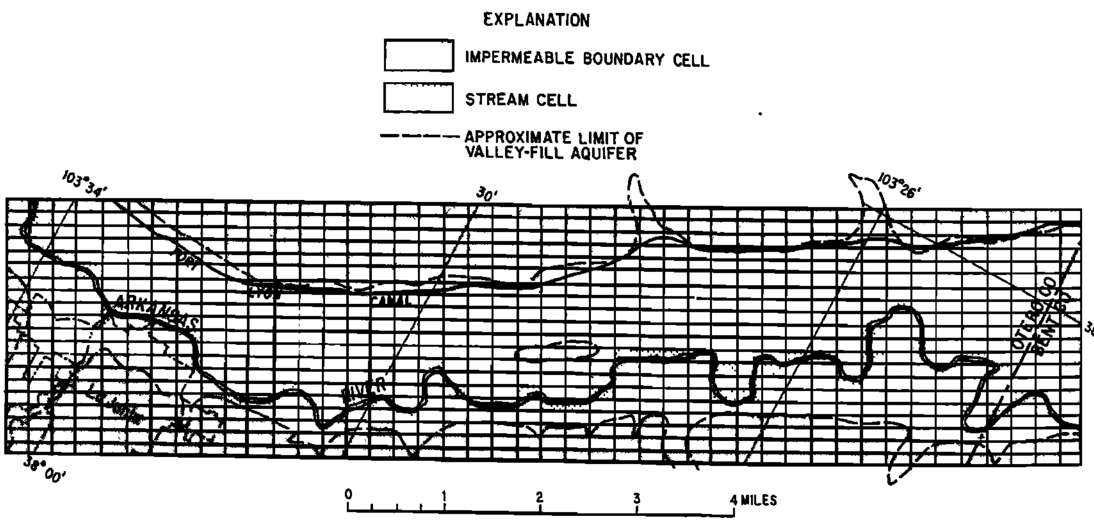

Fig. 2. Finite difference grid used to model study area.

Equation 3 is solved for each node in the finite difference grid. This process is accomplished numerically by using an iterative alternating direction implicit procedure. The derivation and solution of the finite difference equations, the theory of aquifer models, and the advantage of the alternating direction implicit procedure have all been discussed previously in the literature. Some of the more relevant references include *Pinder and Bredehoeft* [1968], *Bredehoeft and Pinder* [1970], *Pinder* [1970], *Prickett and Lonnquist* [1971], and *Rushton and Tomlinson* [1971]. Discussions and examples of the application of digital models to complex stream-aquifer systems are presented by *Bittinger* [1967], *Longenbaugh* [1967], *Pinder and Bredehoeft* [1968], and *Trescott et al.* [1970].

The velocity of groundwater flow must also be known in order to predict the rate and direction of movement of dissolved chemicals. The seepage (pore) velocity at the center of any cell of the finite difference grid is given by

$$v_i = \frac{q_i}{n} = \frac{K_{ij}}{n}\frac{\partial h}{\partial x_j} \tag{4}$$

where v_i is the seepage velocity, L/T; q_i is the specific discharge, L/T; K_{ij} is the hydraulic conductivity of the aquifer, L/T; and n is the effective porosity of the aquifer (dimensionless).

Dispersion equation. If it is assumed that no chemical reactions occur between the water and the aquifer or soil materials that affect the dissolved solid concentration, the equation describing the mass transport and dispersion of dissolved chemical constituents in a saturated porous medium may be written

$$\frac{\partial C}{\partial t} = \frac{\partial}{\partial x_i}\left(D_{ij}\frac{\partial C}{\partial x_j}\right) - \frac{\partial(v_i C)}{\partial x_i} - W_i \tag{5}$$

where C is the mass concentration of dissolved solids, M/L^3; D_{ij} is the coefficient of hydrodynamic dispersion, L^2/T; and W_i is the mass flux of a source or sink, M/L^3T. The theoretical basis and derivation of the dispersion equation are discussed in detail by *Bear et al.* [1968], *Reddell and Sunada* [1970], *Ogata* [1970], and *Bredehoeft and Pinder* [1973]. The first term on the right-hand side of (5) represents the movement of dissolved solids due to hydrodynamic dispersion and is assumed to be proportional to the concentration gradient. The second term describes convective transport, which is proportional to the seepage velocity. The third term represents a fluid source or sink.

Bear et al. [1968] state that hydrodynamic dispersion is the macroscopic outcome of the actual movements of individual tracer particles through the pores and that it includes two processes. One process is mechanical dispersion, which depends upon both the flow of the fluid and the nature of the pore system through which flow takes place. The second process is molecular and ionic diffusion, which because it depends on time, is more significant at low flow velocities. In the field situation existing in the Arkansas River valley the contribution of molecular and ionic diffusion to hydrodynamic dispersion is negligible.

The coefficient of hydrodynamic dispersion is a second-rank tensor. *Scheidegger* [1961] expresses the relationship between the dispersion coefficient, the fluid flow, and the nature of the pore system as

$$D_{ij} = \alpha_{ijmn}\frac{v_m v_n}{|v|} \tag{6}$$

where α_{ijmn} is the dispersivity of the porous medium, L; $v_m v_n$ are the components of the velocity in the m and n directions, L/T; and $|v|$ is the magnitude of the velocity, L/T. He further shows that for an isotropic porous medium the dispersivity tensor can be defined by two constants. These are the longitudinal dispersivity of the medium α_1 and the transverse dispersivity of the medium α_2. These are related to the longitudinal and transverse dispersion coefficients by

$$D_L = \alpha_1 v \tag{7}$$

and

$$D_T = \alpha_2 v \tag{8}$$

respectively. In this analysis it has been assumed that the longitudinal and transverse dispersivities are related by

$$\alpha_2 = 0.3\alpha_1 \tag{9}$$

Although the relationship is somewhat arbitrary, *Bredehoeft and Pinder* [1973] found that it provided reasonable results. The sensitivity of the modeling results to variations in this relationship has not yet been evaluated.

The method of characteristics introduced by *Garder et al.* [1964] is used to solve (5). This technique does not introduce numerical dispersion (artificial dispersion resulting from the numerical process). *Garder et al.* [1964] and *Reddell and Sunada* [1970] have compared solutions obtained by the method of characteristics with those derived by analytical methods and for the cases investigated found good agreement. The development and application of this technique in groundwater problems have been presented by *Pinder and Cooper* [1970], *Reddell and Sunada* [1970], and *Bredehoeft and Pinder* [1973].

The method of characteristics involves placing several moving particles in each cell of the finite difference grid; five particles per cell were used in this study. The location (x and y coordinates) and concentration associated with each particle vary with time. First, each particle is moved a distance proportional to the length of the time increment and the velocity at the location of the particle. The velocity at a point is calculated by bilinear interpolation between adjacent cell nodes, at which the velocities were calculated by using (4). To assure continuity, the maximum distance traveled by a particle is limited so that it will not move more than one half of one cell distance in a given time step. This result is accomplished by adjusting the size of the time increment. After all particles have been moved and particles created or removed as needed at appropriate boundaries, sources, and sinks, the concentration at each node is temporarily assigned the average concentration of all particles then located within that cell. Then the change in concentration due to dispersion is calculated explicitly for each node, and the average concentration at the node is adjusted for this change. The concentration associated with each particle located in that cell is then adjusted for this change.

Representation of the river. Within the study reach the Arkansas River has a sandy channel and is in good hydraulic connection with the aquifer. Stream nodes, shown by a stippled pattern in Figure 2, were designated to represent the river. The meandering river channel necessitated an approximation of its location. At each stream node the stream bed leakage K_z/m was assigned a relatively high value (1.0 ft/s/ft), and the flux between the river and the aquifer was calculated by using (2).

Streamflow was routed downstream by adjusting the total inflow for the flux at each successive stream cell in a downstream direction. At the stream cell at La Junta an additional adjustment was made for the effluent discharged from the La Junta municipal sewage treatment plant. Similarly, an adjustment was made for the flow of the small tributary entering the Arkansas River about 1½ mi downstream from La Junta. The dissolved solid concentration in the river was routed downstream in a like manner, but at La Junta the adjustment also included a correction for the salt load contributed to the river by tributary inflow between the Fort Lyon Canal diversion and the La Junta gaging station. The dissolved solid concentration in the flow leaving a stream cell was calculated from

$$C_{out} = \frac{q_{in}C_{in} + q_f C_f}{q_{in} + q_f} \tag{10}$$

where C_{out} is the dissolved solid concentration in the river leaving a stream cell, M/L^3; C_{in} is the dissolved solid concentration in the river entering a stream cell, M/L^3; C_f is the dissolved solid concentration of the flux between the stream and the aquifer, M/L^3; q_{in} is the streamflow entering a stream cell, L^3/T; and q_f is the flux between the stream and the aquifer (positive for a gaining stream), L^3/T. Thus if the river had been gaining higher-salinity groundwater at a stream cell, then both the flow and the dissolved solid concentration would have increased at that node. If the river were losing water to the aquifer at a stream cell, then the flow leaving the cell should be less than the flow entering the cell. But in this case the dissolved solid concentration in the river would not change ($C_{in} = C_f = C_{out}$).

Summary of calculation procedure. The steps followed in the calculation procedure are outlined in Figure 3. First, the hydraulic properties, boundary conditions, and initial conditions at the start of the simulation period for the study area must be described. Then all hydrologic and chemical stresses imposed on the system during 1 month are defined. Consideration of all data and stresses on a monthly basis is sufficient to describe adequately water level and water quality changes occurring in this stream-aquifer system. But because the accuracy of the numerical calculations is strongly affected by the length of the time period being simulated, several smaller time steps are required to complete each month's simulation.

The model calculates the head distribution in the aquifer at the end of each time step. Hydraulic gradients determined from new water table elevations are used to compute velocities of groundwater flow throughout the aquifer. The changes in concentration due to transport and dispersion of dissolved solids during the time step are computed next on the basis of the groundwater velocity and the concentration gradients. A flow chart outlining the steps taken in this part of the calculation procedure is presented in Figure 4. A short initial time step (10^4 s) is used for each month. The length of each succeeding time increment is progressively increased until the total period of simulation equals 1 month. These steps are then repeated for each month of the simulation period.

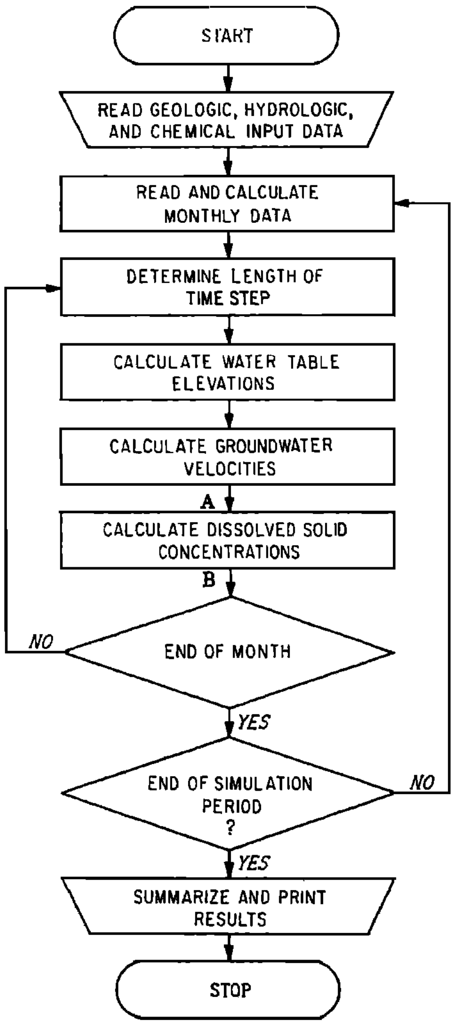

Fig. 3. Simplified flow chart illustrating major steps of calculation procedure. The procedure followed between points A and B is presented in detail in Figure 4.

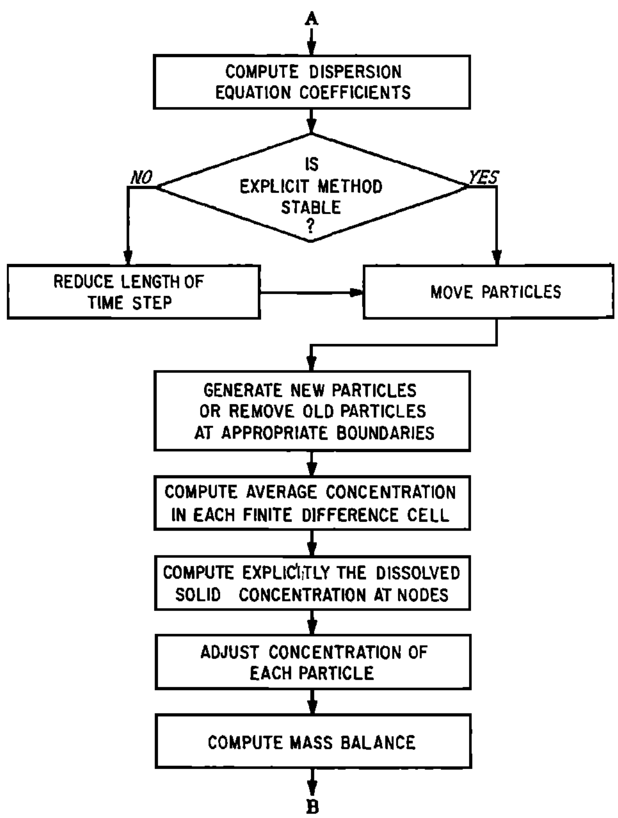

Fig. 4. Flow chart outlining procedure to calculate dissolved solid concentrations (continuity with main flow chart indicated by points A and B).

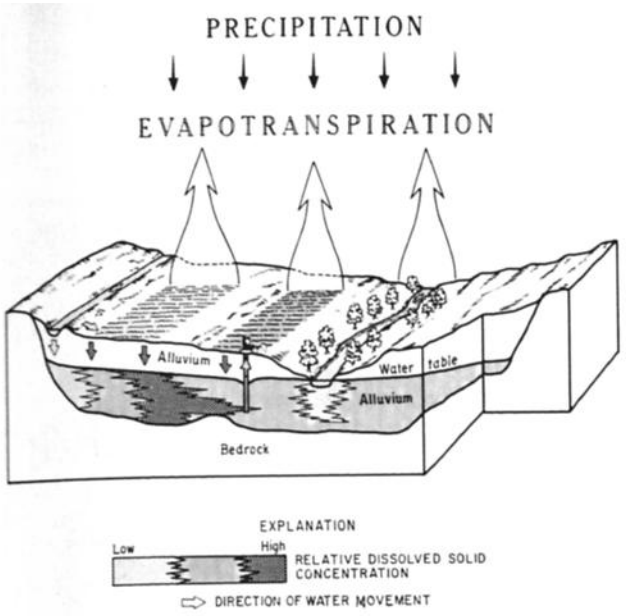

Fig. 5. Idealized block diagram illustrating typical geometry of the stream-aquifer system and the relation between water movement and water quality.

DATA REQUIREMENTS

There are many physical (both natural and controlled) and climatic factors influencing hydraulic and water quality variations in the study reach. Because of the complex interrelations among many of these variables, as illustrated in Figure 5, it is difficult to isolate the relative importance of any single factor. Those factors that were judged to be of only minor significance (such as the effects of temperature and density variations, flow through the unsaturated zone, changes in soil moisture, and chemical reactions between the water and the soil or aquifer materials) were not considered in the model.

Hydrogeologic Data

The floodplain alluvium consists of inhomogeneous deposits of clay, silt, sand, and gravel. The permeability distribution was mapped on the basis of well logs and several aquifer tests. *Hurr* [1966] estimated transmissivities from specific capacity data. Bedrock, consisting mainly of shale with some interbedded limestones, was assumed to be impermeable.

A saturated thickness map was constructed on the basis of the bedrock configuration (determined from well logs) and average annual water table elevations (determined from observation well data). These data were combined with the permeability data to construct a transmissivity map for the aquifer (Figure 6). The average transmissivity within each cell of the finite difference grid was then recorded for the model. Because changes in saturated thickness were small with respect to the total saturated thickness, it was assumed that the transmissivity remained constant over time.

The specific yield of the alluvium was determined from aquifer tests and neutron moisture data to be about 0.20 [*Moore and Wood*, 1967]. This is a common value for this type of deposit and was assumed to be constant throughout the aquifer. No field data were available to determine either the effective porosity or the longitudinal dispersivity of the porous medium. Values for these parameters were selected on the basis of data collected in similar areas. Simulation runs were made by using effective porosity values of 15, 20, and 25% in combination with dispersivity values of 0, 100, and 200 ft. It was found that a uniform effective porosity of 20% produced the best results throughout most of the aquifer, although at some places within the aquifer a value of 15% yielded the best results. A longitudinal dispersivity of 100 ft was also used in the final analysis, but the simulation results were relatively insensitive to adjustments of the dispersivity values.

Observation Well Network

A network of 63 observation wells was maintained during this study. Monthly water samples were collected from 29

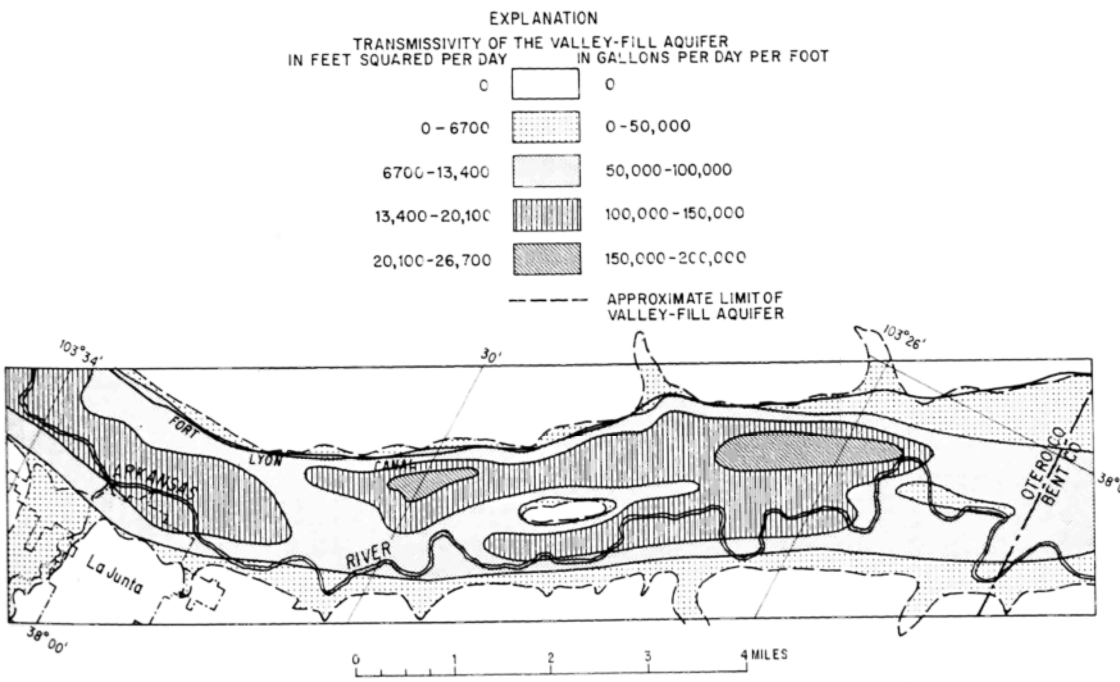

Fig. 6. Transmissivity map of study area [after *Moore and Wood*, 1967].

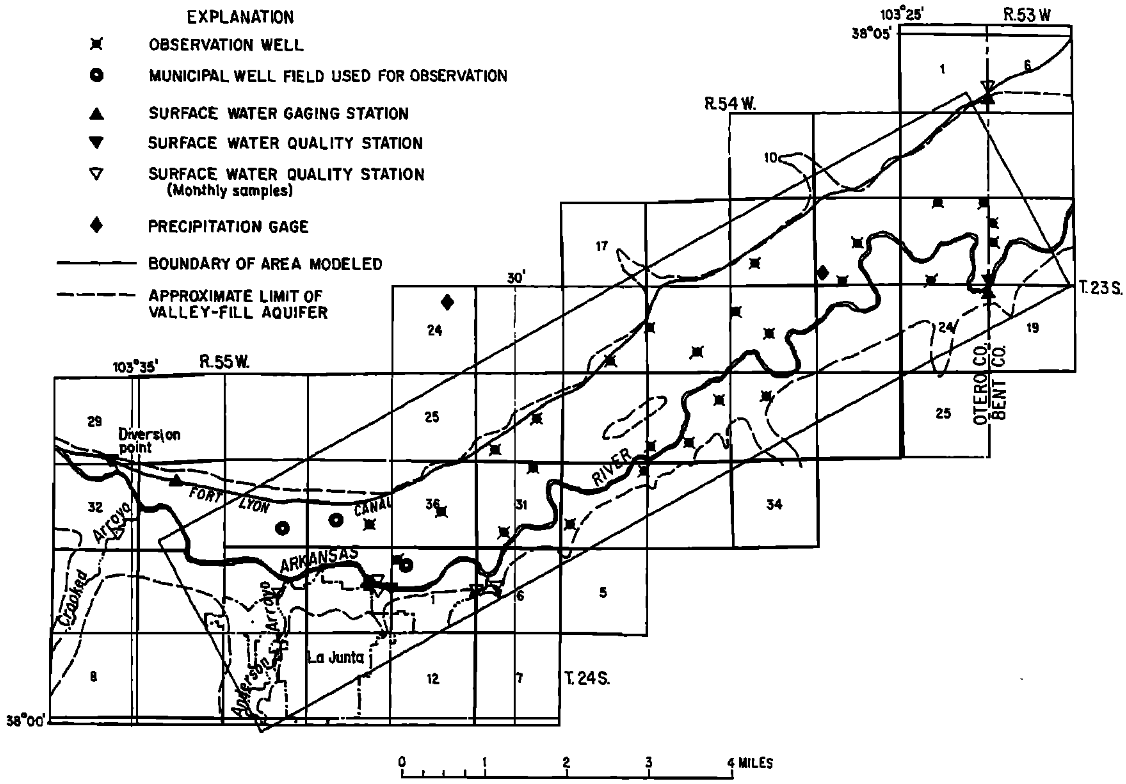

Fig. 7. Data collection network.

wells shown in Figure 7, and water levels were measured in 23 of these. Water levels were also measured semiannually in the 34 additional wells. Observation wells were drilled where they were needed to achieve a monitoring network of sufficient density.

Water levels measured in the observation wells were used to construct a water table elevation map for the beginning of the simulation period (Figure 8). These initial water table elevations were then discretized and used as an initial condition for the simulation. This factor builds into the model realistic initial hydraulic gradients.

However, relatively small errors in the water table elevation data may generate significant errors in the initial hydraulic gradients. The first few computer runs resulted in large discrepancies between the observed and calculated water table elevations in some parts of the study area. The source of this error was determined to be in the input data rather than in the modeling procedure. The land surface elevations of the observation wells had been estimated only from topographic maps, and an instrument survey soon showed that such estimates were commonly in error by a few feet and were inadequate for modeling purposes.

Hydrologic Data

Arkansas River. Gaging stations are located on the Arkansas River near each end of the study reach. Figure 9 shows the mean monthly discharges measured at each gage. Effluent from the La Junta municipal sewage treatment plant

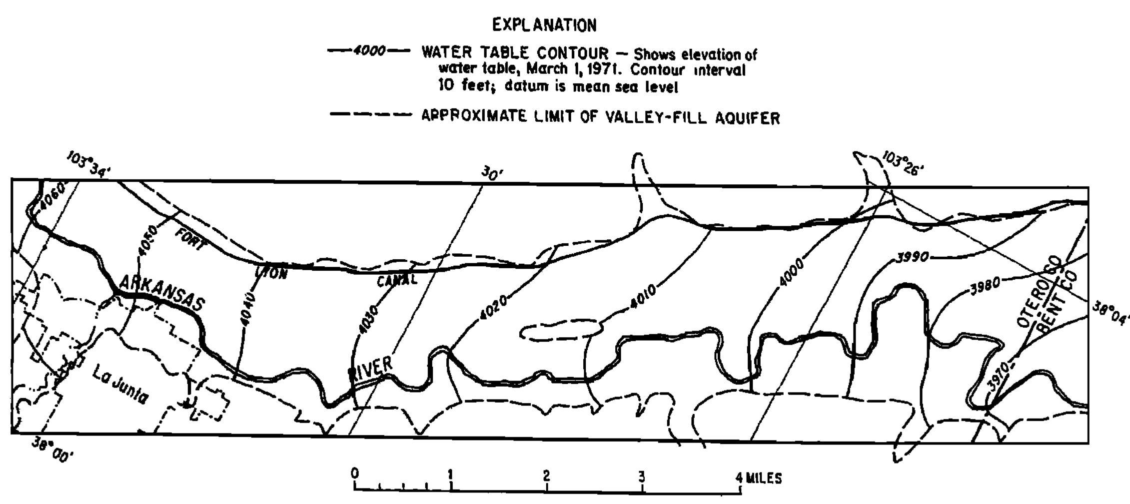

Fig. 8. Water table configuration at beginning of study period, March 1, 1971.

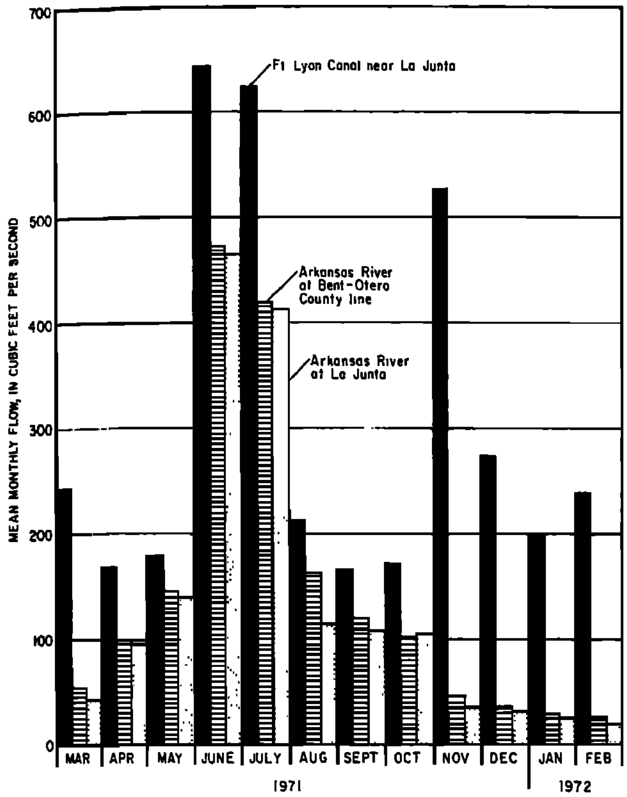

Fig. 9. Mean monthly flows of Arkansas River and Fort Lyon Canal during study period.

enters the river about 500 feet downstream from the La Junta gaging station. This discharge averaged 1.1 ft³/s. Two small tributaries, which join and enter the Arkansas River from the south 1½ mi downstream from the La Junta gaging station, have a combined flow that averaged 0.2 ft³/s.

Changes in river stage affect groundwater flow into a stream, water table elevations, and bank storage [*Cooper and Rorabaugh*, 1963]. The observed stage of the Arkansas River at La Junta fluctuated over a range of about 2 ft during the study period, and Figure 10 shows how the observed stage was approximated by a simple step function in the model.

Fort Lyon Canal. Water is diverted into the Fort Lyon Canal from the Arkansas River west (upstream) of La Junta. From 55 to 95% of the monthly flows of the river above the diversion point was diverted to the canal. The canal is unlined and situated along the northern boundary of the alluvial aquifer.

Within the study reach, canal water is used for irrigation. Diversions from the canal to individual fields are made through approximately 50 small head gates; most of these are controlled by the Fort Lyon Canal Company and measured with small Parshall flumes.

Water table contours and stream-gaging data indicate that some leakage occurs from the Fort Lyon Canal into the aquifer. At times when there were no diversions from the canal, measurements indicated seepage losses as high as 12 ft³/s within the study reach. In the model the canal was treated as a stream having a semipermeable bed. The result was calculated seepage losses ranging from 9 to 14 ft³/s, which are in good agreement with observed seepage losses.

Precipitation. The monthly precipitation was averaged from two gages located in and adjacent to the study area. One gage is located at the La Junta FAA Airport. Its records are published by the *U.S. Department of Commerce* [1971, 1972]. The second gage is an unofficial station maintained by the National Park Service at the Bent's Old Fort National Historical Site. The mean monthly precipitation was assumed to occur uniformly over the entire study reach at a constant rate throughout each month. Precipitation data are summarized in Figure 11.

Potential evapotranspiration. Potential evapotranspiration and consumptive use were estimated on a monthly basis by using a modified *Blaney and Criddle* [1962] equation:

$$u = kf \tag{11}$$

where u is the monthly consumptive use in inches, k is an empirical consumptive use crop coefficient, and f is a monthly consumptive use factor. On the basis of data reported for other areas of similar climate and crop types [*Blaney and Criddle*, 1962] it was assumed that k varied as a function of time from a low of 0.3 in January to a high of 1.1 in July. The value of f is determined from

$$f = tp/100 \tag{12}$$

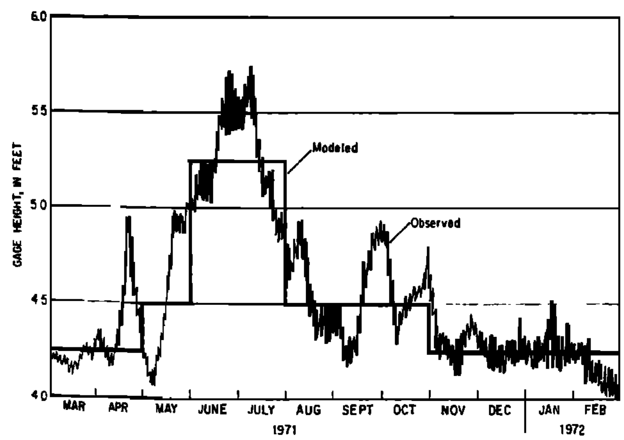

Fig. 10. Observed gage height on Arkansas River at La Junta during study period and step function used to simulate stream stage in model.

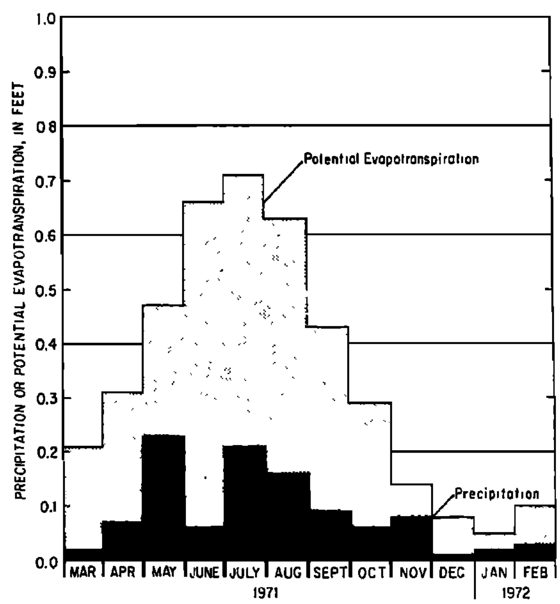

Fig. 11. Monthly average precipitation and potential evapotranspiration, March 1971 to February 1972.

where t is the mean temperature in degrees Fahrenheit and p is the monthly percentage of daytime hours of the year. Monthly values of potential evapotranspiration and consumptive use thus calculated are presented in Figure 11, and supporting data are summarized in Table 1.

Groundwater use. In the study reach, groundwater is used for irrigation and municipal and domestic water supply. There are 68 large-capacity irrigation wells in the area that yield as much as 500–700 gal./min each. Most of these wells are located on the north side of the river. The city of La Junta also maintains three well fields for municipal water supply.

Pumpage from irrigation wells was estimated from total power consumption during the study period on the basis of pump horsepower, lift, and discharge. Monthly pumpage data were not available for wells within the study reach but were available for 24 irrigation wells located about 10 mi east of the study area. Monthly pumping rates for irrigation wells within the area were estimated by assuming that their monthly percentages of the total pumpage equaled the respective average monthly percentages of total pumpage for the 24 wells outside the study area. It was also assumed in the model that all groundwater pumped by irrigation wells was applied uniformly over the land area of the cell in which the wells are located and on the two adjacent cells in a north-south direction, this pattern of irrigation corresponding with the one in the valley. Total groundwater pumpage is shown in Figure 12.

Evapotranspiration from phreatophytes represents a natural groundwater use and in the study area occurs primarily in a narrow strip of land paralleling the river. In the model, phreatophytes were assumed present at all stream cells. The rate of phreatophyte withdrawal was taken to equal the potential monthly evapotranspiration. Groundwater evapotranspiration throughout the remainder of the model area was assumed to be insignificant because the depth to water generally exceeded 10 ft. Because of the limited extent of phreatophytes in this area it was also assumed that phreatophytes did not affect the quality of water in the aquifer. Although this assumption is not strictly valid, there is evidence that some phreatophytes, especially salt cedar, will withdraw both groundwater and dissolved solids contained in the groundwater from an aquifer and during the growing season exude highly saline water on the fronds of the plants [*Gatewood et al.*, 1950, p. 80].

Water quality. Two continuously recording specific conductance meters were located on the Arkansas River; one was above the diversion to the Fort Lyon Canal, and the other was at the downstream gaging station at the Bent-Otero county line (Figure 7). Specific conductance is a measure of the ability of water to conduct an electrical current. Because the specific conductance is related to the number of ions in solution, it can be used to estimate the total dissolved solid concentration of the water. Figure 13 shows that the specific conductance is consistently higher at the downstream end of the study reach.

Analyses of major ions were performed on samples collected at the beginning and end of the study period; the results of the analyses are summarized in Table 2. During the remainder of the period, only the specific conductance was measured. The chemical analyses were used to determine graphically the relationship between specific conductance and dissolved solid concentration (Figure 14). The regression equation for this relation was used to compute the dissolved solid concentration for all water quality samples.

The water quality observed in the aquifer at the beginning of the simulation period is shown in Figure 15. In some areas the dissolved solid concentration of the groundwater changes markedly over a short distance. For example, in the center of the study area the dissolved solid concentration increased from less than 1000 mg/l at the northern boundary to more

TABLE 1. Calculation of Potential Evapotranspiration Using the *Blaney and Criddle* [1962] Equation

	t, °F	p at 38°N	f	k	u inches	u feet
		1971 Data				
March	43.0	8.34	3.58	0.70	2.51	0.21
April	52.6	8.90	4.68	0.85	3.98	0.33
May	60.2	9.92	5.97	0.95	5.67	0.47
June	76.2	9.95	7.58	1.05	7.96	0.66
July	77.1	10.10	7.79	1.10	8.57	0.71
August	75.5	9.47	7.15	1.05	7.51	0.63
September	64.4	8.38	5.40	0.95	5.13	0.43
October	55.4	7.80	4.32	0.80	3.46	0.29
November	42.0	6.82	2.86	0.60	1.72	0.14
December	33.9	6.66	2.26	0.40	0.90	0.08
		1972 Data				
January	30.6	6.87	2.10	0.30	0.63	0.05
February	34.0	6.79	2.31	0.50	1.15	0.10

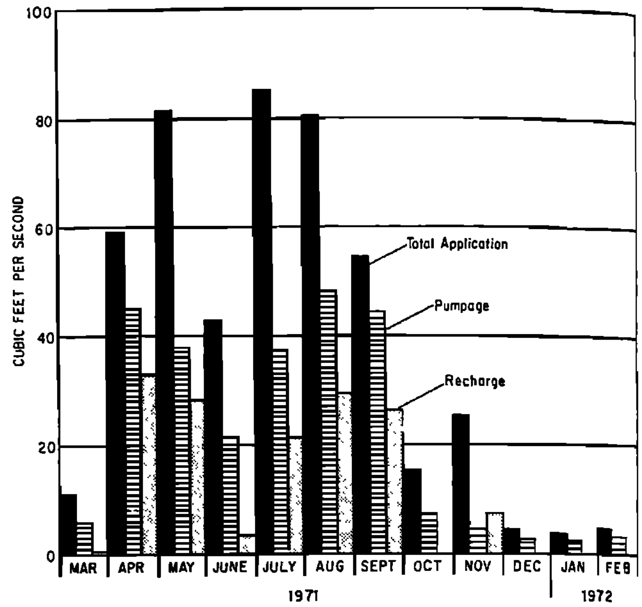

Fig. 12. Monthly total application (irrigation plus precipitation), groundwater pumpage, and calculated recharge to the aquifer within study reach.

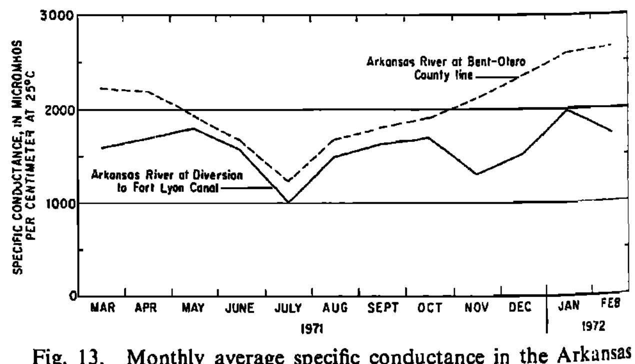

Fig. 13. Monthly average specific conductance in the Arkansas River at the diversion point to the Fort Lyon Canal and at the Bent-Otero county line.

TABLE 2. Summary of Chemical Analyses of 55 Groundwater Samples

Constituent	Maximum	Minimum	Mean
Manganese (Mn)	4.90	0	0.68
Silica (SiO_2)	28.0	14.0	18.9
Iron (Fe)	0.52	0.01	0.06
Orthophosphate (PO_4)	0.49	0	0.04
Calcium (Ca)	480	110	283
Magnesium (Mg)	220	39	102
Potassium (K)	9.2	1.7	5.5
Sodium (Na)	590	83	254
Bicarbonate (HCO_3)	497	126	294
Chloride (Cl)	150	27	74
Fluoride (F)	2.5	0.6	1.1
Sulfate (SO_4)	2300	430	1310
Nitrite + Nitrate (as N)	25	0	2.7
Dissolved solids (calculated)	3850	805	2200
Total hardness	1900	450	1130
Specific conductance, micromhos at 25°C	4400	1150	2700
*p*H	8.3	7.1	
Water temperature, °C	17.5	4.5	13.4
Sodium adsorption ratio	6.7	1.3	3.2

Values are concentrations in milligrams per liter except for specific conductance, *p*H, temperature, and sodium adsorption ratio. All samples were collected during February 1971 or February 1972.

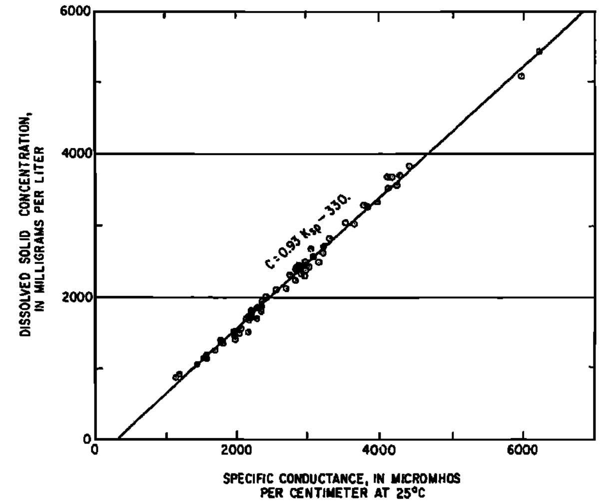

Fig. 14. Relationship between specific conductance K_{sp} and dissolved solids concentration C in study area.

than 3000 mg/l near the center of the valley about ½ mi away. The direction of groundwater flow, irrigation practices, and the location of the river seem to be among the major factors influencing the observed water quality pattern.

Water samples were collected biweekly from the Arkansas River at La Junta. These samples showed that about one half of the total increase in the dissolved solid concentration observed in the river between the Fort Lyon Canal diversion and the Bent-Otero county line was due to sources upstream from the La Junta gaging station, primarily Anderson Arroyo and Crooked Arroyo (Figure 7). Much of the flow of Crooked Arroyo consists of return flow from irrigated land southeast of La Junta. The biweekly samples from the Arkansas River at La Junta and the continuous record collected at the Fort Lyon Canal diversion were correlated to obtain an estimate of the monthly average increases in dissolved solids attributable to sources upstream from La Junta. In the model this increase was introduced at the La Junta gaging station.

Water samples were also collected each month from several tributaries in the study reach. The effluent from the municipal sewage treatment plant at La Junta had an average dissolved solid concentration of 5200 mg/l. The minor tributary located about 1½ mi downstream from the La Junta gage had an average dissolved solid concentration of 6500 mg/l.

Applied water. In the study area, water is applied to the land surface from three sources: natural precipitation, surface water irrigation, and groundwater irrigation. Each source is of different quality. Precipitation was assumed to have no dissolved solids. Irrigation water diverted from the canal was assumed to have the monthly mean dissolved solid concentration recorded at the upstream surface water quality station. Water applied from irrigation wells was assumed to be at the dissolved solid concentration in the aquifer observed or calculated at the beginning of each month at the specific location of each irrigation well. The dissolved solid concentration of the total applied water at each point was then considered to equal the average concentration derived from mixing the three sources of applied water.

More than 50% of the study area is irrigated with water from the Fort Lyon Canal (Figure 16). Irrigation water is normally applied to a field through ditches and furrows. The surface water application rates were generally lowest in the western part of the study area (near La Junta) and highest in the eastern part of the study area. In each month the surface

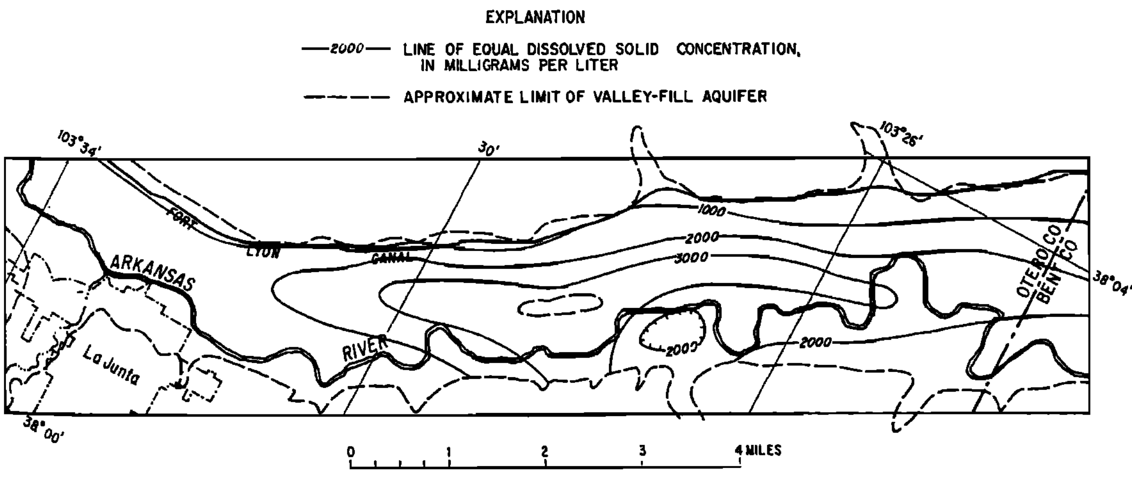

Fig. 15. Dissolved solid concentration in aquifer at beginning of study period, March 1, 1971.

TABLE 3. Surface Water Applications

	Subarea 1, ft	Subarea 2, ft	Subarea 3, ft
	1971 Data		
March	0.032	0.063	0.118
April	0.076	0.091	0.125
May	0.111	0.124	0.124
June	0.156	0.295	0.306
July	0.127	0.259	0.344
August	0.071	0.134	0.231
September	0.029	0.030	0.022
October	0.041	0.026	0.064
November	0.039	0.056	0.225
December	0	0	0.080
	1972 Data		
January	0	0	0
February	0	0	0

water application was assumed to be uniform within each of the three subareas outlined in Figure 16; the average monthly applications within each subarea are presented in Table 3.

Groundwater recharge. Parts of the total applied water are evaporated, consumed by crops, stored in the soil, and recharged to the aquifer by percolation through the soil. The amount of recharge must be considered in the model, but recharge is probably the most difficult factor to evaluate in the hydrologic system. It varies in time and space as a complex function of many physical and climatological variables.

Because changes in soil moisture storage are negligible over extended periods of time, the monthly recharge for each cell was estimated as a function of total applied water and potential evapotranspiration. Equations to calculate recharge that have the following properties were derived (R. R. Luckey, written communication, 1971):

1. The ratio of an increment of recharge to an increment of applied water equals 1 when the total applied water exceeds the potential evapotranspiration.

2. This same ratio is less than 1 when the total applied water is less than the potential evapotranspiration.

3. When the total applied water is less than the potential evapotranspiration, recharge increases as total application approaches potential evapotranspiration.

Within these constraints the recharge during a given time interval and in a given area may be expressed in terms of derivatives as

$$dR/dA = 1 \qquad A \geq E \qquad (13)$$

$$dR/dA = (A/E)^i = W^i \qquad A \leq E \qquad (14)$$

where R is the total recharge, L; A is the total applied water (irrigation plus precipitation), L; E is the potential evapotranspiration, L; i is a recharge parameter greater than zero (dimensionless); and $W = (A/E)$ is the normalized applied water.

Integrating (13) and (14) between zero and A permits recharge to be expressed explicitly as

$$R = E\left(\frac{1}{i+1} - 1\right) + A \qquad A \geq E \qquad (15)$$

$$R = \frac{A^{i+1}}{E^i(i+1)} \qquad A \leq E \qquad (16)$$

The fraction of the total applied water that is recharged may be obtained by dividing (15) and (16) by A to yield

$$R_f = \frac{1}{W}\left(\frac{1}{i+1} - 1\right) + 1 \qquad W \geq 1 \qquad (17)$$

and

$$R_f = \frac{W^i}{i+1}. \qquad W \leq 1 \qquad (18)$$

respectively, where $R_f = (R/A)$ is the recharge fraction. Figure 17 shows the relation between the recharge fraction and normalized applied water for different values of the recharge parameter i. Note that as the value of the recharge parameter becomes very large, the recharge function approaches the limit at which applied water up to potential evapotranspiration is consumed and applied water greater than potential evapotranspiration is recharged.

It was assumed that the recharge rate did not exceed the infiltration capacity of the soil. However, field inspections did show that some flooding of fields occurred occasionally. Because most of the surface runoff from the fields was returned directly to the river, recharge to the aquifer was slightly overestimated.

The recharge parameter is a constant that represents the integrated effects of several physical, climatic, and model characteristics. The permeability and moisture content of the soil, the type and density of plant or crop cover, and the uniformity of application within each finite area and over time are among the factors that affect the value of the infiltration parameter. Although sufficient data are not presently

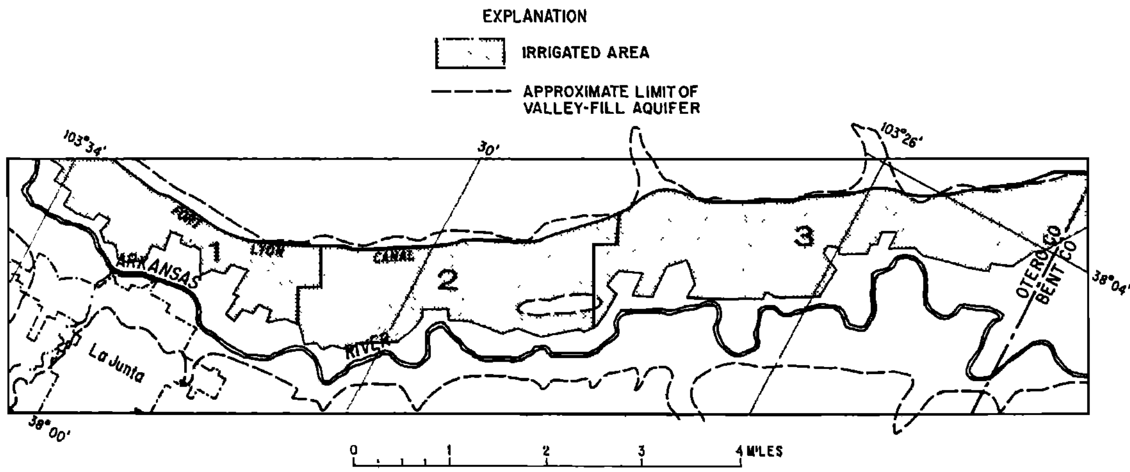

Fig. 16. Study reach showing three subareas irrigated with water from the Fort Lyon Canal.

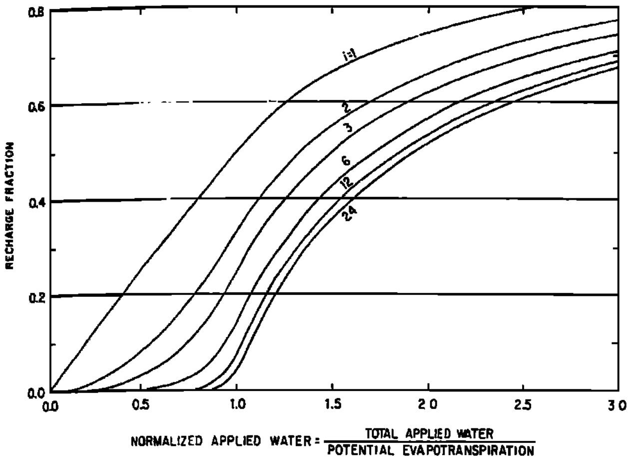

Fig. 17. Relationship between recharge fraction (fraction of total applied water that is recharged) and ratio of total applied water to potential evapotranspiration for selected values of the recharge parameter i (after R. R. Luckey, written communication, 1971).

available to determine the relationship between the recharge parameter and any other factor, for this study area a value of 12 for the recharge parameter yielded recharge rates that produced the best agreement between observed and calculated water level changes. However, because recharge is relatively insensitive to increasing the recharge parameter above a value of 12 and because there are errors and uncertainties in other hydrologic input data, in retrospect it may have been sufficient for this investigation to have assumed simply that applied water up to potential evapotranspiration is consumed and applied water greater than potential evapotranspiration is recharged.

The total recharge within the study reach was calculated to average about 32% of the total application during the 1-yr study period. The recharge percentage of total application varied from less than 1% to greater than 55% on a monthly basis. During any 1 month the cell-to-cell variation of the recharge fraction was even greater.

The dissolved solid concentration of the recharge water was calculated by assuming that the total mass of dissolved solids in the recharge water is the same as that in the total applied water. The increase in concentration in the recharge water is then simply proportional to the decrease in volume of water due to evapotranspiration losses.

Underflow. Groundwater underflow occurs through the alluvium across the upper and lower boundaries of the modeled study reach. Darcy's law was used to estimate the quantity flowing through each of these boundary cells on the basis of the hydraulic gradients existing at the start of the simulation period. The total underflow entering the study reach across the upstream boundary was calculated to be 2.1 ft^3/s, and the total underflow leaving the study reach through the downstream boundary was calculated to be 0.9 ft^3/s. These rates were assumed to remain constant during the study period. Groundwater interchange between the alluvium and the adjacent bedrock is probably insignificant owing to the low permeability of the bedrock. Thus the valley walls and the bedrock surface beneath the alluvium were treated as impermeable boundaries.

SIMULATION RESULTS AND MODEL VERIFICATION

Water table fluctuations. Water table elevations throughout the study area were computed for the end of each month in the simulation period. These elevations were compared with field observations to calibrate and verify the model. At any given time the range in water table elevations within the study reach was approximately 100 feet. Figure 18 shows that the observed and computed water table configurations after 1 yr (March 1, 1972) agree quite well.

Only a small annual change was noted from the March 1, 1971, to the March 1, 1972, water table. However, there were significant seasonal water level fluctuations observed in many observation wells during the study period. The measured water levels in 23 observation wells varied during the study period by an average of about 3 ft. An estimate of the accuracy of the model based on these 23 observation wells indicates that within the study reach the water table elevation was predicted within 1 ft of the observed value more than 90% of the time. The two largest sources of error were probably the monthly pumpage data and recharge rates. Figure 19 shows the observed and computed water levels at two observation wells. Part of the discrepancy between the observed and computed water levels is also due to the fact that the loca-

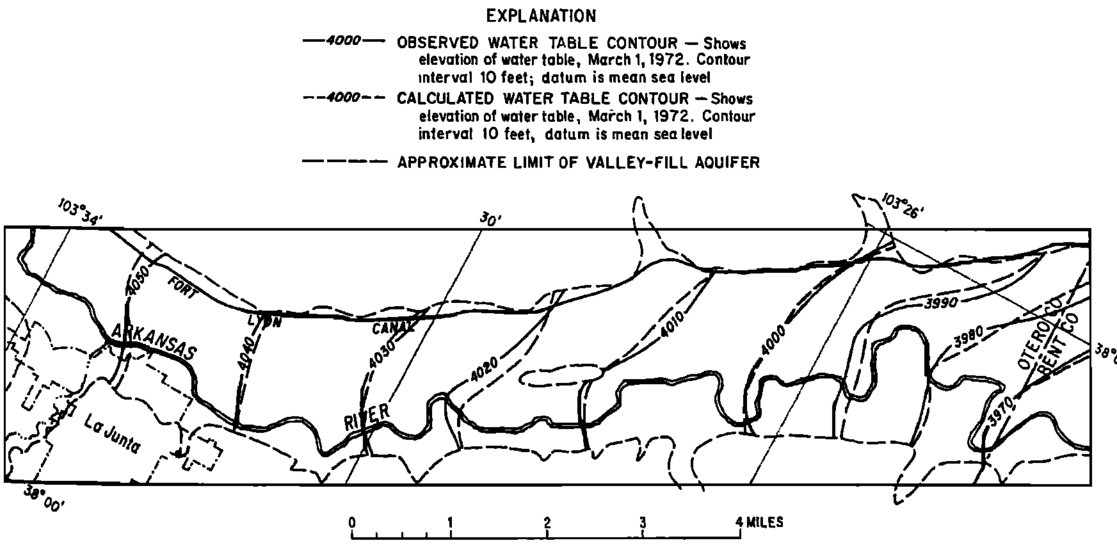

Fig. 18. Observed and calculated water table configurations at the end of study period, March 1, 1972.

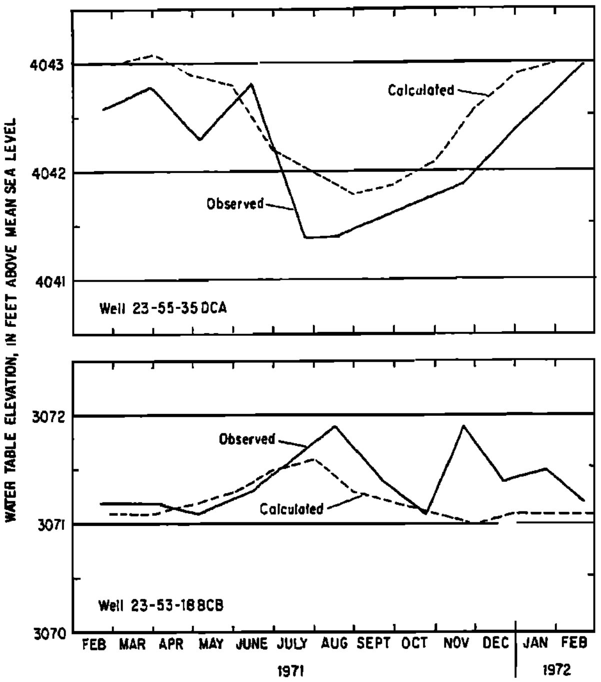

Fig. 19. Observed and calculated water levels in two wells during study period.

tion of each observation well does not coincide with the respective nodal location of the finite difference cell. The importance of this factor for any observation well is indicated by the difference between the observed and calculated initial water table elevations. The well location numbers indicate the township, range, section, and position within the section, in that order. The system of numbering wells is described in more detail by *Major et al.* [1970].

Groundwater quality. The digital computer model was also used for calculating the dissolved solid concentration throughout the aquifer at the end of each month. Figures 20 and 21 present the observed and calculated dissolved solid concentrations, respectively, in the aquifer on March 1, 1972. Also shown are areas in which the dissolved solid concentration had changed by more than 10% since March 1, 1971. The overall water quality patterns of the observed and calculated data are in fairly good agreement. However, there are several areas where discrepancies appear.

Increases in dissolved solid concentration were observed along the northern and eastern boundaries of the study area, whereas decreases were noted at places in the center of the valley. Figures 20 and 21 show that calculated changes are in relatively good agreement with the observed changes. Observed and calculated variations of the dissolved solid concentration in the aquifer during the study period at two wells are shown in Figure 22. These data indicate to the authors that the model can successfully reproduce observed changes in dissolved solid concentration in the aquifer.

An assessment of the accuracy of the model can only be performed in light of any inherent variability in the field data. That is, the variability at an observation point or within the area of a finite difference cell must be evaluated before the significance of the variation between observation points or between finite difference cells can be interpreted properly. To help evaluate the former, samples were collected simultaneously from each of the five wells located in the west municipal well field. Because this well field encompasses an area equivalent to about one fourth of the area of one finite difference cell and no well is located farther than about 250 ft from the center of the well field, it was assumed that the variability of specific conductances of these five samples would provide a measure of the chemical variation within small representative volumes of the aquifer and that none of the variation was due to the existence of a larger-scale concentration gradient across the well field. The specific conductances of these samples had a mean of 2110 μmho/cm at 25°C and a standard deviation of 208 μmho/cm at 25°C, which is about 10% of the mean. Although the variation observed within this one well field is not necessarily representative of the entire study area, it does indicate that the criteria used to evaluate the accuracy of the model should not be less than ±10% of the observed data.

A comparison between the observed data from 29 observation wells and the computed data for the corresponding closest finite difference cells indicates that the dissolved solid

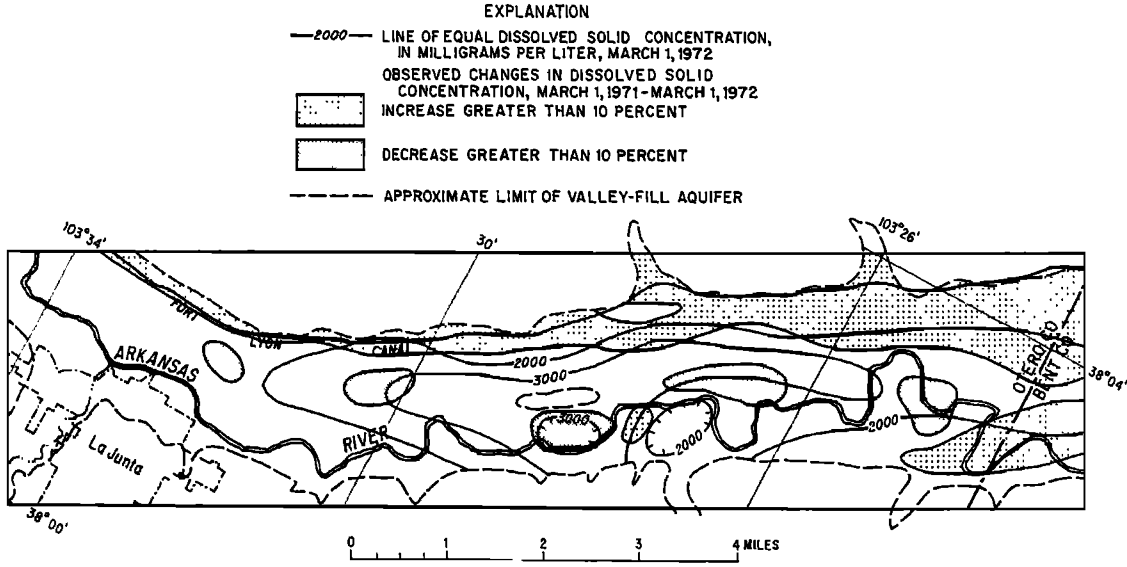

Fig. 20. Observed dissolved solid concentration in aquifer on March 1, 1972, and areas of observed changes since March 1, 1971.

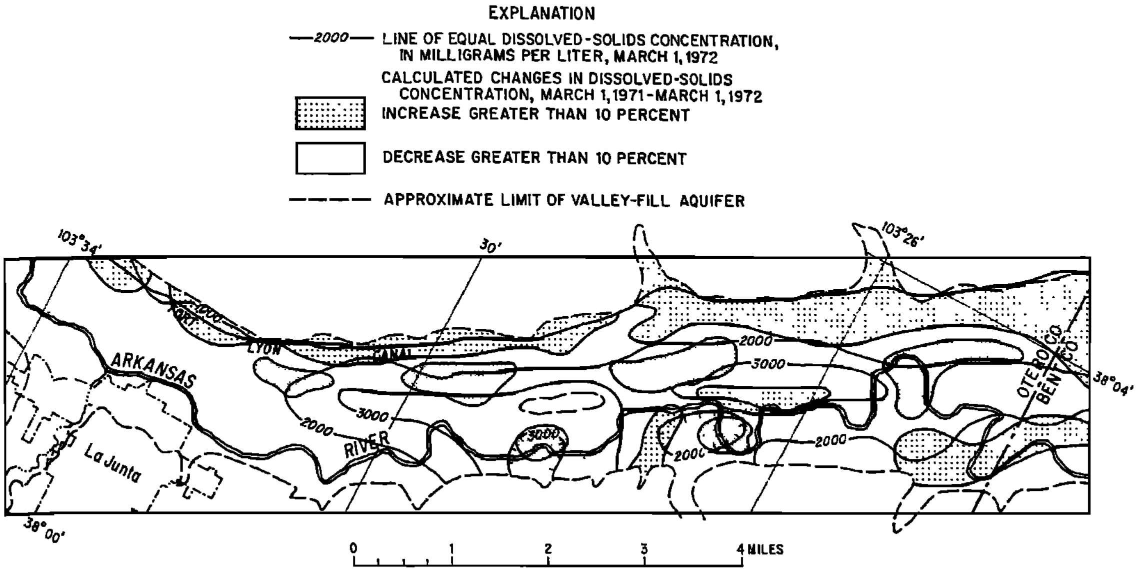

Fig. 21. Calculated dissolved solid concentration in aquifer on March 1, 1972, and areas of calculated changes since March 1, 1971.

concentration was reproduced within 10% of the observed value approximately 80% of the time. The average deviation was 200 mg/l. The differences are probably related to both data deficiencies and modeling errors. One such source of error is the time difference between the date of determination for the calculated data, which was the end of each month, and the actual date of sample collection for the observed data. Also in areas where the concentration gradient is steep a small error in locating the lateral position of a line of equal dissolved solid concentration will induce a large error in the dissolved solid concentration at a given point.

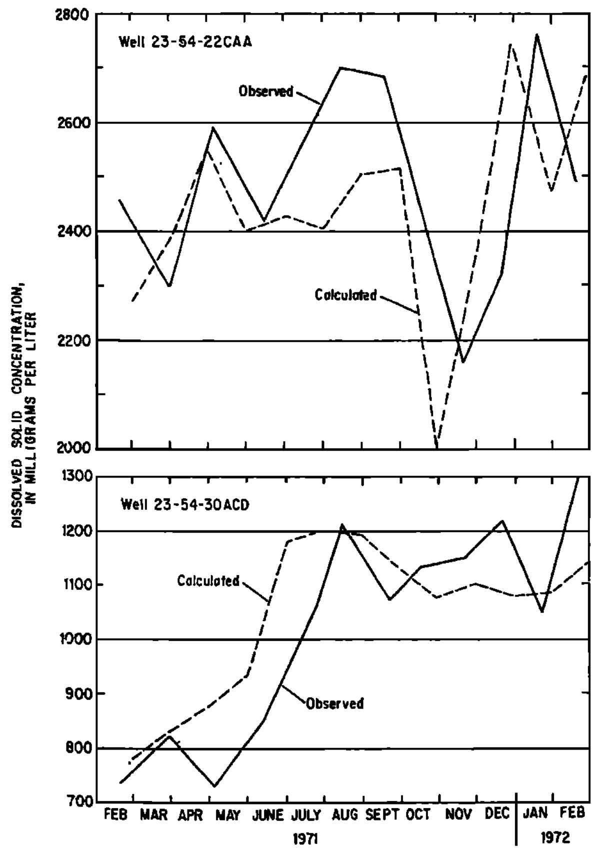

Fig. 22. Observed and calculated dissolved solid concentrations of groundwater at two wells during study period.

Streamflow. The net change in the observed flow of the Arkansas River between the upstream and downstream gages was usually only a small percentage of the actual discharge and was initially assumed to be the result of the interchange of water between the stream and the aquifer. The reliability of the simulation model is in part measured by how accurately these stream gains and losses are predicted. The stream gains and losses varied with time, primarily in response to pumping, irrigation practices, and fluctuations of river stage.

In using the observed data, which represent the difference in flow between the two gaging stations, several factors must be considered. First, tributary inflow was unmeasured and may have been significant at times during the spring and summer when thunderstorm activity was the greatest. Thus it is likely that some of the observed gains do not represent groundwater flow into the stream. For example, during August 1971, when the greatest gain in streamflow occurs (49 ft^3/s), an analysis of the daily hydrographs indicates that 40 ft^3/s could be attributed to tributary inflow occurring in two distinct flood events. During a later field inspection of these tributaries, high water marks were found that indicated recent flooding. Second, measurement accuracy may be no better than 2% for monthly mean flows (C. T. Jenkins, oral communication, 1971). The error in the difference in flow between the two gaging stations may then be as large as the sum of the errors at each. Thus during periods of high flow in the spring and summer the measurement error in the observed difference in flow may be larger than the actual gain or loss. Third, direct return to the river of excess irrigation water applied to the fields was observed to occur through drainage ditches. During one field inspection the total direct return flow to the river within the study reach was estimated to be between 4 and 6 ft^3/s.

The importance of each of these three factors would be minimal during the winter. Figure 23, which presents a comparison between the observed and calculated net changes in streamflow, shows that the best agreement occurs during the winter, when streamflow is low, tributary flooding is not

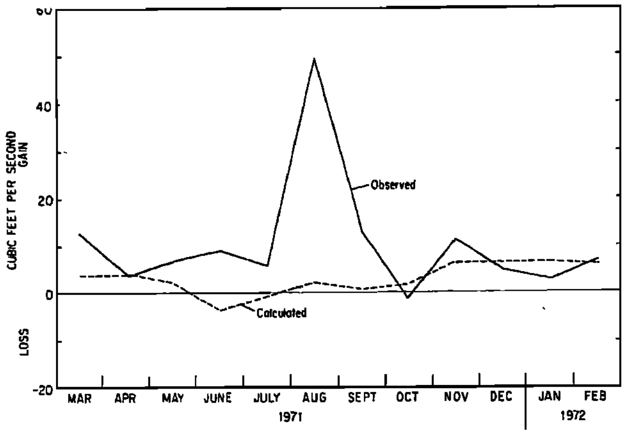

Fig. 23. Observed and calculated changes in mean monthly flow of Arkansas River between La Junta and the Bent-Otero county line.

likely, and irrigation is minimal. This suggests that the calculated streamflow gains and losses may be more representative of the actual interchange of water between the stream and the aquifer than the observed differences in flow between the two gaging stations.

The model indicated that the river was gaining water in the reach during most months with an average computed gain of approximately 3 ft²/s. Gains were greatest during the fall and winter and least during the spring.

Quality of surface water. The dissolved solid concentration in the Arkansas River tends to be lowest during periods of high flow because of the low dissolved solid content of snowmelt water and precipitation runoff. Groundwater of higher dissolved solid concentration that discharges to the river within the study reach causes an increase in the salinity of the river. The monthly average changes in surface water quality within the study reach are presented in Figure 24. Although groundwater discharge to the river was relatively constant, the magnitude of the change in concentration of dissolved solids in the river was also affected by the flow of the river because of dilution effects. The observed increase in the

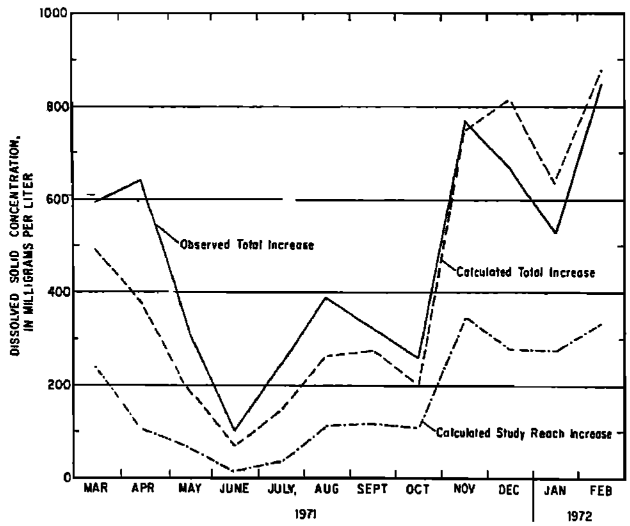

Fig. 24. Observed and calculated monthly average changes in dissolved solid concentration in the Arkansas River. The total increase represents the change between the Fort Lyon Canal diversion and the Bent-Otero county line, and the study reach increase occurs between La Junta and the Bent-Otero county line.

dissolved solid concentration in the river between the Fort Lyon Canal diversion and the Bent-Otero county line averaged 475 mg/l. This is an increase of about 40% over the 1175-ml/l average value observed at La Junta. The greatest increases occurred during the winter, and the smallest increases occurred during late spring and early summer.

An average increase of 220 mg/l in the dissolved solid concentration in the Arkansas River was calculated to occur within the study reach (from La Junta to the Bent-Otero county line) and represents that part of the total increase attributable to irrigation return flows that occur via the groundwater system. When the dissolved solid load contributed by tributary inflow is accounted for, the observed and calculated total increases (between the Fort Lyon Canal diversion and the Bent-Otero county line) agree quite well.

Downstream variations. Stream gains and losses vary from point to point down the river in response to pumpage and recharge in adjacent reaches of the aquifer. Thus at any one time the rate of change of dissolved solid concentration in a downstream direction will be affected by the location of gaining stream segments and the quality of groundwater adjacent to those gaining stream segments.

Downstream changes in streamflow and water quality calculated by the model for August 1971 are typical and are illustrated in Figure 25. The changes in streamflow were plotted by using a three-point moving average of calculated data for successive stream cells. The change in concentration noted at point B represents adjustments for the salt load contributed by upstream tributary inflow and the effluent from the La Junta municipal sewage treatment plant, and the change at point C represents an adjustment for the minor tributary entering the Arkansas River at that location. The greatest and most consistent stream losses are noted in the up-

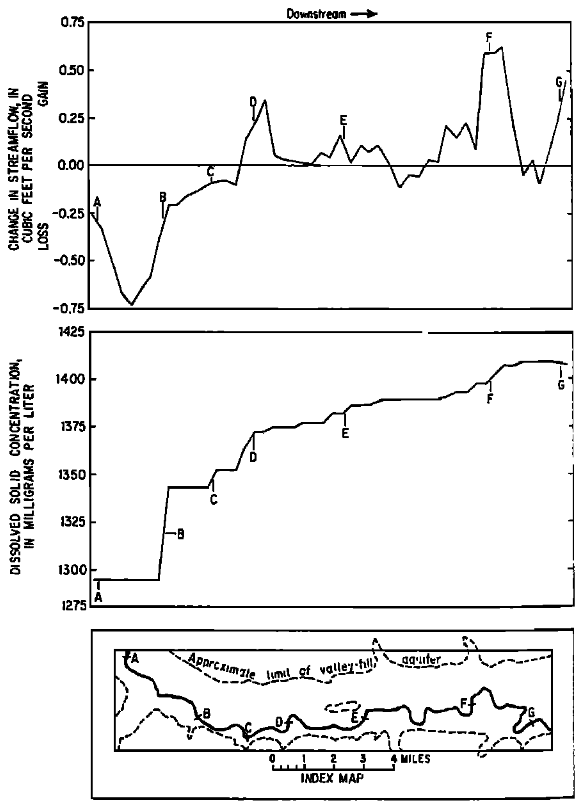

Fig. 25. Calculated downstream changes in streamflow and dissolved solid concentration in Arkansas River, August 1971.

per one third of the study reach (points A–C). In this part of the study reach, large withdrawals of groundwater occur at the La Junta municipal well fields. Also recharge is slight because irrigation is less in this part of the study area. The greatest stream gains were computed to occur at three sites (D, E, and F) where because of meandering the river flows for short distances perpendicular to the general down-valley direction. The highest rates of increase of the dissolved solid concentration in the stream were associated with reaches with these large gains.

RECOMMENDATIONS FOR FUTURE STUDY

Several factors known to affect the flow or dissolved solid concentration of groundwater were not considered in this model because their importance was believed to be negligible in this study area. These include the effects of water temperature and density variations, flow through the unsaturated zone, changes in soil moisture, and chemical reactions between the water and the soil or aquifer materials. However, some of these factors might be significant in other areas, and methods or equations to evaluate them should be incorporated into the general simulation model; the U.S. Geological Survey is currently attempting to incorporate chemical reactions into the model. Within the study area the effects of adsorption and ion exchange would have to be evaluated in order to predict the behavior of most specific ions.

Improved or new techniques are needed to determine field values of certain aquifer properties, such as effective porosity and dispersion coefficients, which now cannot readily be evaluated. In some areas the spatial and temporal variations of these parameters may be important.

CONCLUSIONS

1. Complex water quality variations exist in an irrigated stream-aquifer system. These can be predicted with a digital simulation model that couples a finite difference technique, used to solve the flow equations, with the method of characteristics, used to solve the transport (dispersion) equations.

2. The model described was successfully applied to an 11-mi reach of the Arkansas River valley in southeastern Colorado for a 1-yr period, although the need for some further refinements was indicated. The simulation results greatly aided the understanding of water quality variations within this area and their relation to irrigation practices. Changes in the dissolved solid concentration of groundwater were predominantly controlled by convective transport and by mixing with recharged water of different quality. In this simulation the influence of hydrodynamic dispersion was relatively minor.

3. Many input data are required for the model, and the reliability of the modeling results is affected by the accuracy of these data. A sensitivity analysis helps in the definition of accuracy requirements for each of the input parameters. This is particularly useful for those parameters that cannot readily be evaluated in the field. Because the relative magnitude of each hydrologic factor can vary from one locality to another, so can the respective accuracy requirements.

4. The model can be applied to other irrigated areas where the required data are available or can be collected. With few programing modifications the technique can be used for many problems in which it is desired to predict the rates and directions of movement of contaminants through a saturated porous media.

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